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Title:

Scale-dependency of effective hydraulic conductivity on fire-affected hillslopes

Date:

2016-07-01

Citation:

Langhans, C., Lane, P. N. J., Nyman, P., Noske, P. J., Cawson, J. G., Oono, A. & Sheridan, G. J. (2016). Scale-dependency of effective hydraulic conductivity on fire-affected hillslopes. *Water Resources Research*, 52 (7), pp.5041-5055. <https://doi.org/10.1002/2016WR018998>.

Persistent Link:

<https://hdl.handle.net/11343/291915>

Title: Scale-dependency of effective hydraulic conductivity on fire-affected hillslopes

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Key points:

- Effective hydraulic conductivity parameters after fire were compared between scales
- Small-scale methods resulted in 4.8 times larger estimates than large-scale methods
- Runoff-, ponding- and K_s -distributions are currently not modeled well enough

This is the author manuscript accepted for publication and has undergone full peer review but has not been through the copyediting, typesetting, pagination and proofreading process, which may lead to differences between this version and the [Version record](#). Please cite this article as [doi:10.1002/2016WR018998](https://doi.org/10.1002/2016WR018998).

20 Abstract:

21 Effective hydraulic conductivity (K_e) for Hortonian overland flow modeling has been defined
22 as a function of rainfall intensity and runoff infiltration assuming a distribution of saturated
23 hydraulic conductivities (K_s). But surface boundary condition during infiltration and its
24 interactions with the distribution of K_s are not well represented in models. As a result, the
25 mean value of the K_s distribution ($\overline{K_s}$), which is the central parameter for K_e , varies between
26 scales. Here we quantify this discrepancy with a large infiltration data set comprising four
27 different methods and scales from fire-affected hillslopes in SE Australia using a relatively
28 simple yet widely used conceptual model of K_e . Ponded disk (0.002 m²) and ring
29 infiltrometers (0.07 m²) were used at the small scales and rainfall simulations (3 m²) and
30 small catchments (ca 3000 m²) at the larger scales. We compared $\overline{K_s}$ between methods
31 measured at the same time and place. Disk and ring infiltrometer measurements had on
32 average 4.8 times higher values of $\overline{K_s}$ than rainfall simulations and catchment scale estimates.
33 Furthermore, the distribution of K_s was not clearly log-normal and scale-independent, as
34 supposed in the conceptual model. In our interpretation water repellency and preferential
35 flow paths increase the variance of the measured distribution of K_s and bias ponding towards
36 areas of very low K_s during rainfall simulations and small catchment runoff events while
37 areas with high preferential flow capacity remain water supply limited more than the
38 conceptual model of K_e predicts. The study highlights problems in the current theory of
39 scaling runoff generation.

40 Index Terms: 1838, 1847, 1850

41 Keywords:

42 Effective hydraulic conductivity, infiltration, scale-dependence, runoff modeling, wildfire,
43 Australia

1. Introduction

Traditional infiltration theory based on Darcy type flow relies on the assumption that soil structure is *homogeneous* over some distance [e.g. *Green and Ampt*, 1911]. Saturated hydraulic conductivity (K_s) is one of the most important parameters in this conceptual model. Instead of K_s , the name of effective hydraulic conductivity (K_e) has been used when applying the original infiltration theory to naturally *heterogeneous* soils. K_e values vary in time due to changing soil properties or rainfall intensity [*Kidwell et al.*, 1997; *Langhans et al.*, 2010; *Nearing et al.*, 1996; *Paige et al.*, 2002; *Risse et al.*, 1995; *Ruan et al.*, 2001; *Van den Putte et al.*, 2013]. These studies were largely empirical and concerned with *a priori* estimation of parameter values for soil erosion prediction models such as WEPP [*Nearing et al.*, 1989].

Earlier, in hydrology, K_e signified a parameter with values that are invariant within a specified space-time domain: it was a specific type of mean of a distribution, e.g. the arithmetic, harmonic or geometric mean or some intermediate values (as reviewed in: *Binley et al.* [1989]; *Blösch and Sivapalan* [1995]). This definition had the problem that, while acknowledging spatial heterogeneity, it implicitly assumed unlimited water supply, which is unrealistic during most natural rainfall events. *Hawkins* [1982] conceptualized rainfall intensity as a water supply limitation for area-averaged steady-state infiltration rates. This idea has been subsequently employed to re-define K_e as area-averaged infiltration rate at field-saturated conditions for a given rainfall intensity [*Smith and Goodrich*, 2000]. In this definition, K_e is not an effective parameter in the sense of a constant, but a function of water supply. This conceptual model of K_e , not always in name but in purpose, has been extended to account for effects of changing surface boundary conditions such as water depth and water supply rate, during Hortonian overland flow [*Dunne et al.*, 1991; *Karssenbergs*, 2006; *Langhans et al.*, 2013; *Morbidelli et al.*, 2006]. The application of this definition of K_e to calculate area-averaged infiltration rates (f_e) with traditional infiltration theory is not always

straightforward, but ensemble equations for f_e have been formulated [Langhans *et al.*, 2014; Morbidelli *et al.*, 2006; Smith and Goodrich, 2000].

Regardless of which definitions for K_e one adheres to, when using K_e in models of Hortonian overland flow, one aims at integrating actual heterogeneity and flow processes present within a model unit. For practical relevance the values of K_e , or the distribution parameters they depend on, need to be such that observations of runoff match predictions. Perhaps unsurprisingly, it has proven difficult to find unique values or functions. Not only are K_e values very sensitive to rainfall intensity, antecedent moisture conditions, and changing soil surface conditions during an event [Binley *et al.*, 1989; Risse *et al.*, 1995], but they are also scale-dependent. Natural soils on hillslopes have a complex macro- and microscopic structure, and there is inadequate or incomplete theory that links the most important flow processes, such as matrix- and macropore flow, and surface runoff [e.g. Beven and Germann, 1982; 2013]. Scale-dependency of K_e stems from the non-linearity of the flow processes and their interactions with a complex, often spatially correlated soil and surface structures that are not well represented by the homogeneity assumption of Darcian flow [Beven, 1989; Grayson *et al.*, 1992].

Boundary conditions in this study are defined as flux and head variables at the surface, such as water depth, ponding extent (inundation of the microtopography), and water supply rates (runon). These boundary conditions vary in time and space, within and between measurement methods and scales. One particular aspect of the scale-problem of K_e is the actual versus the modeled surface boundary condition and distribution of K_s . The interaction of surface boundary conditions with the distribution of K_s and their effect on infiltration rates at different scales are largely unresolved in models; it is an important, and generally neglected sub-problem of the ‘non-linearity of flow processes’ issue. While in traditional infiltration theory, the surface boundary was simply ponded or not ponded, subsequent models have

introduced some sophistication. Rainfall intensity-dependency of K_e was introduced by assuming a distribution of K_s where the portion of the soil surface where rainfall intensity was larger than K_s was ponded, and the remainder not ponded [Hawkins, 1982; Wood *et al.*, 1986]. Later developments have introduced the concept of ‘runon infiltration’ where the surface can be ponded not only by rainfall excess, but also runoff water from upslope [Corradini *et al.*, 1998; Jones *et al.*, 2013; Karssenberg, 2006; Morbidelli *et al.*, 2006]. Water depth increases pressure at the surface and is therefore directly linked to infiltration rates at a location [e.g. Reynolds and Elrick, 1990]. But this fact has not yet been included in models of K_e , where water depth is a distribution within the micro-topography. Some researchers however have linked the water depth distribution to the distribution of K_s and therefore K_e , but have not considered water repellency [Dunne *et al.*, 1991; Langhans *et al.*, 2014]. Others have focused on modeling the supply of surface water to surface-connected macro-pores [Léonard *et al.*, 2001; Ruan and Illangasekare, 1998; Weiler, 2005].

The problem is that all these models have been developed with particular environments in mind, where the studied boundary condition was assumed to be important. But none of the studies has actually isolated the effect of surface boundary conditions by measurement under natural rainfall-runoff conditions, which would be very difficult. It is also not clear if the studied effects are indeed mathematically well presented, because different conceptualizations and their calibrated parameter values can result in comparable predictions; this is the problem of model identifiability [Kirchner, 2006; McDonnell *et al.*, 2007]. As a consequence there is no generally accepted theory that can explain observed scale-effects in a broad range of conditions. A related problem is that infiltration measurements at different scales necessarily require different measurement methods that all have different boundary conditions. Only a ‘perfect’ model would result in estimates of K_e that are independent of the method.

Studies addressing the surface runoff scaling problem mostly propose new models with a 'better' representation of boundary conditions than a previous model that it seeks to supplant, and evaluate the performance with measurements at different scales (see section before last). This approach is a valuable type of study to advance our understanding of a system, and improve predictions. Here we choose a different approach, where we seek to quantify the overall importance of the boundary conditions (given by scale and method) in the estimation of K_e . We have chosen a model of K_e with a relatively simple representation of surface boundary conditions, and which can be applied at all scales and for all methods. In order to control for other factors than surface boundary conditions affecting values of K_e , methods are compared with data at the same location and time. Resulting differences in K_e using the same conceptual model can be attributed to the *aggregate effect* of all aspects of *surface boundary conditions* particular to scale and method that the model does not account for. As far as we know, this approach is new.

The magnitude of the scale- and method effect will be different for each study area. We chose burnt hillslopes, because this environment has not yet been extensively studied with regards to infiltration on a variety of scales. However, for a burnt area, Ebel et al. [2012] compared values of field saturated hydraulic conductivity (K_{fs}) between techniques at the small core scale, and Nyman et al. [2014a] and Sheridan et al. [2007] found contrasting values of K_{fs} between the rainfall simulation and mini-disk or ring infiltrometer scales also for burnt areas. Where present, we included data from unburnt areas at those burnt sites. We used a unique data set of infiltration measurements comprising four methods (disk- and ring infiltrometers, rainfall simulations, and small catchment runoff measurements) within the South East Australian Ecosystem of fire-affected forests. Good understanding and models of runoff generation in fire-affected forests are critical for studies of post-fire erosion, water quality, and geomorphic hazards. Site-specific factors are not part of the focus of this paper, but on

fire-affected hillslopes we can expect distributions and processes different from traditional assumptions in infiltration theory [Ebel and Moody, 2013], which makes this environment an important testing ground for the generality of existing models.

The underlying purpose or motivation of the present work is to critically test an existing way to predict large scale surface runoff from fire-affected forest hillslopes using smaller-scale infiltration measurements. The work should help identify areas where theoretical development of surface-runoff scaling is required. The specific objective of this study is to quantify the overall importance of the boundary conditions, given by different scales and measurement methods, in estimating the central parameter of K_e . For this we present a large data set of infiltration measurements from fire-affected areas, and analyze them using a well-used infiltration model.

2. Reference model for K_e

2.1. Log-normally distributed Strips Concept (LSC)

We chose a model that is 1) widely used, 2) consistent with a widely used definition of K_e , and 3) with a relatively simple representation of the surface boundary condition to allow quantifying the aggregate effect of boundary conditions between scales and methods. The model in question was first presented by Smith et al. [1990] and Smith and Goodrich [2000], and will be called hereafter ‘Log-normally distributed Strips Concept’ (LSC). The LSC assumes a log-normal distribution of K_s where they are arranged in strips downslope, which means that there is no interaction between runoff and infiltration on the runoff strips. The conceptual model was taken up in the runoff and erosion model KINEROS [Smith et al., 1990] and KINEROS2 [Goodrich et al., 2006]. Smith and Goodrich [2000] have developed

an empirical formulation for f_e and K_e based on this idea. Here we are interested in K_e ($L\ T^{-1}$), which is relevant for rainfall simulations at apparent steady state (Equations 11a and 11b of Smith and Goodrich [2000]):

$$\frac{K_e}{K_s} = \left[1 + \left(\frac{K_s}{i} \right)^p \right]^{-1/p} \quad (1a)$$

where

$$p = \frac{1.8}{CV^{0.85}} \quad (1b)$$

and where i is rainfall intensity ($L\ T^{-1}$), $\overline{K_s}$ ($L\ T^{-1}$) is the mean of the log-normal distribution of K_s ($\overline{K_s} = \exp(\mu + 0.5\sigma^2)$), CV (-) the coefficient of variation of the log-normal distribution of K_s ($CV = \sqrt{\exp(\sigma^2) - 1}$), with μ the mean and σ^2 variance of the log-transformed values. In words, K_e increases more strongly with rainfall intensity when CV is higher, and it is equal to K_s when CV is 0 (Figure 1).

[Figure 1]

The assumption of a log-normal distribution of K_s has been justified by ring infiltrometer measurements [Baker and Bouma, 1976; Sharma et al., 1980]. This observation has been widely repeated [Haws et al., 2004; Lai and Ren, 2007; Lauren et al., 1988; Mallants et al., 1997; Mertens et al., 2002; Sisson and Wierenga, 1981; Wuest, 2005]. KINEROS2 contains the same conceptual model of infiltration as Equation 1 a and b, only for the time-dependent case (f_e , not shown), published in Smith and Goodrich [2000]. A steady rainfall modeled in

KINEROS2 would therefore yield final infiltration rates equal to K_e . In the remainder ‘LSC’ refers to both KINEROS2 and Equation 1 a) and b), as both are based on the same theory.

KINEROS2 allows for some runoff infiltration broadly defined as infiltration of runoff further downslope: after-flow infiltration (infiltration after rainfall excess has ceased) and runoff from a model unit (‘plane’ in KINEROS2) with low infiltration onto a plane with high infiltration are possible, but no runoff due to spatial variability within the model unit is represented. KINEROS2 is popular for erosion and runoff studies [e.g. *Al-Qurashi et al.*, 2008; *Kalin et al.*, 2003], including post-wildfire runoff [*Canfield et al.*, 2005; *Chen et al.*, 2013; *Goodrich et al.*, 2005; *Langhans et al.*, 2016; *Sidman et al.*, 2015]. The LSC accounts only for rainfall intensity and not for water depth, variable water supply by runoff, or other complex surface boundary conditions. The LSC is therefore suitable for quantifying the importance of the aggregate of these effects across scales and methods.

2.2. Terminology

We can now clarify the relationship between K_e , $\overline{K_S}$, K_s and K_{fs} . In this study we are interested in K_e , because as the apparent (effective) hydraulic conductivity its meaning is the same for all infiltration models (definitions) and at all scales (see introduction). But for quantitative comparison, we chose a specific definition of K_e (LSC, previous section) for which the comparable parameter across scales and methods is $\overline{K_S}$. If $\overline{K_S}$ is scale-dependent, then so is K_e , only that we cannot easily compare the scale-dependency of K_e , because K_e also depends on rainfall intensity (i.e. water supply rate at the surface).

K_s is used only in the context of distribution, because in the LSC K_s does not have any specified spatial dimension. For the purpose of quantifying this distribution each $\overline{K_S}$ value of

the small scale measurements was interpreted as an instance of K_s from which the distribution can be assessed.

Measurements of saturated hydraulic conductivity in the field are often labeled field-saturated hydraulic conductivity (K_{fs}), because some areas of the soil matrix remain unsaturated, especially in the presence of water repellency [Nyman *et al.*, 2010]. During all types of measurements presented here the soil was field-saturated rather than saturated, but for our purpose, K_{fs} was interpreted as $\overline{K_S}$.

3. Study area

The study sites are located in the North-Eastern and Central Highlands in Victoria, SE Australia (Figure 2). The area is mountainous with an elevation up to 1800 m, and steep slopes of up to 40°, but typically between 25 and 35°. It is mostly covered by forests dominated by the Eucalypt genus (*eucalyptus*). In areas that receive more annual precipitation (1200 – 2000 mm), mostly at mid-and higher elevation, forests are ‘Wet’, which means that they are not water-limited. At lower elevations the forest is a ‘Dry’ mixed eucalypt species forest (Wet and Dry are designations used in the Victorian Ecological Vegetation Classification, [DSE, 2012]). There is a strong aspect effect, as well, with equatorial-facing (northerly) slopes having higher levels of net radiation than polar-facing (southerly) slopes, and thus being significantly dryer (see Nyman *et al.* [2011] for a more detailed description of the regional physical environment). The aridity index [Budyko, 1974], which is a measure of long-term water or energy limitation ranges from 3 on rain-shaded (less than 1000 mm annual precipitation) and equatorial-facing slopes (water limited) to below 1 on polar-facing, rainy (up to 2000 mm) slopes (energy limited) [Nyman *et al.*, 2014b]. Aridity index and forest type (wet, dry) are related to soil hydraulic properties and soil depth, with deeper soils and higher

hydraulic conductivities at lower aridity indices, and shallower, stony soils with lower hydraulic conductivities at higher aridity [Sheridan *et al.*, accepted]. The region experiences regular severe forest fires during the months of December to February that spread over ten thousands of hectares. During the first two years after fire, the soil surface layer, underlying a thin hydrophilic layer, is highly water repellent [Nyman *et al.*, 2014a]. In the dryer spectrum the forest soils are less well developed and infiltration rates are low, leading to overland flow and erosion, whereas in the wetter forests well structured soils have relatively high infiltration rates despite the occurrence of water repellency [Nyman *et al.*, 2010; Sheridan *et al.*, 2007].

[Figure 2]

4. Methods

The study draws on a large runoff and infiltration data set collected from burned and unburned forests in SE Australia between 2005 and 2010. The LSC model is applied to calculate the mean value of the K_s distribution ($\overline{K_s}$), which is the central parameter of K_e , for data from the larger scales. By comparing values of different methods at the same locations and times, other factors of infiltration are controlled for such that the importance of surface boundary conditions can be quantified.

4.1. Measurements

Various studies throughout the last two decades have measured infiltration or runoff in the study area using different methods. We chose studies and sites where two different methods with different scales of measurement were carried out within about four weeks of each other. This approach ensured that soil conditions were similar and allowed direct comparison,

controlling for much of seasonal or random site-specific effects. We retained a data set of six sites and four different methods (Table 1) with a total of 628 individual observations. The data have been partly published in international journals [Noske *et al.*, accepted; Nyman *et al.*, 2010; Nyman *et al.*, 2014a; Sheridan *et al.*, 2007], and partly published in theses [Cawson, 2012; Oono, 2010]. For disk infiltrometer (DI) measurements a retention ring of 5.3 cm diameter was inserted 2-3 cm into the soil to avoid lateral loss. A constant ponding depth of 5 mm was maintained while readings were taken in 15 second intervals for 10-15 minutes. If a reading did not change after 5 minutes, the experiment was stopped, which gives this method a lower limit of accuracy of about 3.4 mm h⁻¹. The method is described in more detail Nyman *et al.* [2014a]. For single ring infiltrometer (RI) measurements a ring with a diameter of 30 cm was inserted 5 cm into the soil. A constant water depth of 2 - 60 mm (due to sloping terrain) was maintained at the centre of the ring. Adjustment for water depth in a single ring were made after Reynolds [1993], with more details given in Sheridan *et al.* [2007]. Rainfall simulations (RFS) were conducted on plots of 1.5 x 2 m with a central oscillating boom with a kinetic energy of 295 kJ ha⁻¹ mm⁻¹ for about 30 minutes [Sheridan *et al.*, 2007]. Some rainfall intensity rates were around 60, but most around 100 mm h⁻¹. Runoff rates from natural rainfall from two small catchments (CAT), an equatorial-facing (North) (STCN: 3094 m²) and an adjacent polar-facing (South) (STCS: 2567 m²), were measured with a large tipping bucket where runoff passes a bed-load trap, and enters the buckets with a nominal tipping volume of 12 L [Noske *et al.*, accepted]. More information on the rainfall events is given in the results section.

[Table 1]

These four types of measurement differ in scale, boundary conditions and distribution of individual measurements (Table 2). Boundary conditions are sometimes controlled, such as with unlimited supply of water at a fixed water depth for the disk and ring infiltrometer measurements. We chose 5 mm of ponded head, which was within the natural range of 0 to around 10 mm for rainfall simulations and runoff events [Dunne and Dietrich, 1980; Emmett, 1978; Langhans *et al.*, 2013; Smith *et al.*, 2010]. But for the two larger scales they cannot be controlled, and are difficult to measure. As the data set was collected for different purposes, boundary conditions such as inundation and water depths were not measured for rainfall simulations let alone for catchment runoff. While rainfall simulations had steady supply of water, catchment rainfall is unsteady. These differences, summarized in Table 2, do not constrain the analysis, but they are the object of the study, and they can be used during interpretation of the results.

[Table 2]

4.2. Determination of $\overline{K_S}$ and CV

Disk and ring infiltrometer scale (0.002 - 0.07 m²)

Final infiltration rates during disk and ring infiltrometer experiments (after correction for water depth for RI) were interpreted as $\overline{K_S}$ ($f_e = K_e = \overline{K_S}$), and $\overline{K_S} = K_s$ for the purpose of quantifying the distribution of K_s . $\overline{K_S}$ at this scale is therefore consistent with the LSC model. 61 % of the disk infiltrometer measurements were below the reading threshold 3.4 mm h⁻¹. The actual distribution of these values between 0 and the reading threshold is unknown, but we assumed that their distribution had a mean of half the threshold (1.7 mm h⁻¹). While this

approach is arbitrary, we did not opt to simulate a distribution for the values between 0 and 3.4 mm h⁻¹, because we do not know the real overall distribution.

Rainfall simulation scale (3 m²)

For rainfall simulation data, the apparent steady state infiltration rates were calculated from the last three runoff readings of each rainfall simulation. According to the LSC definition this steady state infiltration rate equals K_e . We solved Equation 1a with known K_e and i for $\overline{K_S}$ using the CV (Equation 1b) of the disk or ring infiltrometer measurements made at the same site and time. Measurements were made on the same hillslopes with separation of less than a few tens of meters.

Catchment scale (3000 m²)

For catchment data $\overline{K_S}$ was derived by calibrating KINEROS2 so that modeled and observed runoff hydrographs had the least possible deviations at one minute time intervals, as measured by the Nash-Sutcliffe Model Efficiency (ME) [Nash and Sutcliffe, 1970]. For three periods of two months (Table 1), centered on the date when disk infiltrometer measurements were made, all storms that produced runoff were included in the calibration exercise, but only modeled runoff events with a ME of more than 0.5 were retained for further analysis. The rainfall events were derived from a single rain gauge that was situated on the ridge separating the two catchments. Rainfall events were defined as separate when there were 30 minute intervals without rain before and after rainfall.

The chosen configuration of KINEROS2 reflects the regional conditions after fire (Table 3).

A two-layer system, where the first layer functions as interception storage, reflects a highly

porous non-repellent surface layer over a sub-surface layer with a high degree of water repellency [Nyman *et al.*, 2014a]. Canopy interception after fire was not taken into account, but was probably small, as most leaves had been consumed during fire. The interception value is thus an estimate for the average storage potential of the non-repellent surface layer, assumed to cover the whole catchment uniformly. Water infiltrating into this layer is partially inhibited from entering the second layer. It was therefore assumed that the soil's capillary drive at the surface (matric potential) was insignificant compared with the gravitational potential. Based on our judgment, reasonable values for microtopography and water temperature were chosen, but they were not measured (Table 3). For more information about these parameters we refer the reader to the online documentation of KINEROS2 (<http://www.tucson.ars.ag.gov/kineros/>) or Smith *et al.* [1995].

[Table 3]

CV for the application in KINEROS2 was based on the variability of disk infiltrometer measurements at the same time and distributed within the catchments. $\overline{K_s}$ and Manning's n ($L^{-1/3} T$) were optimized. Manning's n is related to the unit width discharge (q) as:

$$q = \frac{S^{1/2} h^{5/3}}{n} \quad (3)$$

where S is the slope gradient, and h is water depth. KINEROS2 uses equation 3 with the factor 1.49 in the denominator for English units. Values of Manning's n obtained during

optimization were thus multiplied by this factor to conform to SI units. KINEROS2 uses the kinematic wave equations, which are a simplification of the St. Venant equations, to calculate overland flow distribution in time and space. For optimization, a set search algorithm was applied, where based on reasonable starting values for the two calibration parameters, ME was calculated for a single storm. Subsequently, every parameter combination of half the starting values' step size in each direction of the parameter space from the starting value was used to calculate a set of ME. Retaining the parameter combination that yielded the highest ME as new starting values, reducing the step size, and repeating the process allowed the set algorithm to converge on an optimized pair of Manning's n and $\overline{K_S}$. Optimized parameter combinations were found for each storm that produced sufficient runoff, and for these storms we assumed that the objective function surface (values of ME across the parameter space) was smooth enough for this algorithm to yield unique parameter values, which seems reasonable as a similar optimization study yielded unique values [Langhans *et al.*, 2014].

Distribution of K_S at all scales

The observed distribution of log-transformed $\overline{K_S}$ values of all methods was tested with the Lilliefors test [Lilliefors, 1967]. The null hypothesis of log-normal distribution was rejected for DI and CAT at a significance level of 0.05 (Table 2). The null hypothesis for RI and RFS could not be rejected, but p-values were relatively low, which casts doubt on the log-normal distribution hypothesis ($p = 0.27$ and $p = 0.06$, respectively). As a consequence $\overline{K_S}$ was calculated as the arithmetic mean, and CV as the standard deviation over the mean, rather than using mean and CV definition of a log-normal distribution (section 2.1).

4.3. Analysis

The observed distributions of $\overline{K_s}$ of all measurement scales were compared. Mean values of $\overline{K_s}$ and mean values of CV were calculated for each period of measurement at a site. If the period in question contained data from burnt or unburnt parts of the site, they were compared separately. The Wilcoxon ranking test [Wilcoxon, 1945] for significant differences between medians of two populations was applied to separate periods and sites. We compared $\overline{K_s}$ across scales and measurement methods: ratios of $\overline{K_s}$ and CV were calculated, where values of the smaller scale measurements were divided by the larger scale measurement:

$$\overline{K_s} \text{ ratio: } \frac{\text{Mean } \overline{K_s} (CV) \text{ of small scale (DI or RI)}}{\text{Mean } \overline{K_s} (CV) \text{ of larger scale (RFS or CAT)}} \quad (4a)$$

$$CV \text{ ratio: } \frac{CV \text{ of small scale (DI or RI)}}{CV \text{ of larger scale (RFS or CAT)}} \quad (4b)$$

The distribution of K_s in the LSC is assumed to be log-normally distributed and scale-independent, which means that both CV and $\overline{K_s}$ would be the same at all scales if boundary conditions of the measurements were fully accounted for in the LSC. Deviations of $\overline{K_s}$ and CV -ratios from 1 can thus be attributed to the combined effect of actual vs. assumed distribution and boundary conditions.

The LSC does not account for runoff infiltration directly. The related model of Hawkins and Cundy [1987] predicts that with an arrangement of higher K_s further downslope K_e will equal $\overline{K_s}$. In the LSC we have also calculated $\overline{K_s}$ for rainfall simulations and catchment data with $CV = 0$ (no runoff possible), and compared it with $\overline{K_s}$ using CV from the disk and ring infiltrometer measurements in order to quantify the effect of runoff within the LSC.

395

396 5. Results

397 During three periods, and in the two catchments (STCN and STCS), a total of 137 runoff
398 events was recorded, some of which resulted in only a few tips of the tipping buckets (Table
399 5). For about half of the events parameters were calibrated with an ME of more than 0.5, and
400 an average of 0.86. The remaining events had too low ME values, or the algorithm could not
401 converge. ME values were mainly too low for events with low runoff totals, and an
402 unresolvable hydrograph due to sparsity of tips. Mean optimized $\overline{K_S}$ ranged from 1.5 to 5.0
403 mm h^{-1} , and mean Manning's n from 0.14 to 0.23, with consistently higher values for both
404 parameters in the southern aspect catchment (Table 4). The 77 runoff events that were finally
405 used in the calibration differed between the three measurement periods (Table 5). Rainfall
406 events were on average about 150 minutes in the first two periods, and 420 minutes in the
407 third period. Rainfall totals between time periods and the two catchments ranged from an
408 average of 6.8 to 20.7 mm, and the average one-minute maximum rainfall intensity of the
409 events was between 18.4 and 37 mm h^{-1} , with the most intensive rainfall in March 2010.
410 Runoff coefficients ranged from 0.26 to 0.39, with higher runoff coefficients on the
411 equatorial-facing aspect, which is consistent with the lower $\overline{K_S}$ values.

412

413 [Table 4, Table 5]

414

415 The distribution of $\overline{K_S}$ of RI, RFS and CAT appear relatively straight on the log-normal
416 probability plot (Figure 3), but log-normality was either rejected or uncertain using the

Lilliefors test (Table 2). Values of disk infiltrometer measurements above the reading threshold (39 % of total) are not log-normally distributed either.

[Figure 3]

Mean $\overline{K_S}$ values of the smaller scale disk and ring infiltrometer measurements were on average 4.8 times larger than values estimated for rainfall simulations and catchments (Figure 4, Table 6). However, only some of the differences were significant, due to high variability and sometimes low numbers (Table 6). The $\overline{K_S}$ ratio was highest with a mean of 8.0 where ring infiltrometer measurements were compared to rainfall simulation values, second highest with a mean of 3.8 for the comparison of disk infiltrometer over rainfall simulation measurements, and lowest with a mean of 2.6 for disk infiltrometer over catchment values. The CV was on average 4.7 times higher for smaller than for the larger scales, with a similar pattern as the $\overline{K_S}$ ratio between sites and methods (Table 6).

[Table 6, Figure 4]

There are some recognizable trends that are important for the interpretation of the results. First, there is a clear signal of aridity, with lower aridity sites having $\overline{K_S}$ values that are lower by an order of magnitude for DI and RFS (Table 6). Secondly, whereas $\overline{K_S}$ values are not clearly lower for burnt versus unburnt sites, the $\overline{K_S}$ ratio itself is somewhat higher for burnt sites.

6. Discussion

6.1. Distribution of K_s

The distribution of K_s was not clearly log-normal at any scale. While for ring infiltrometer measurements and rainfall simulations the log-normal assumption was reasonable, it was not log-normal in the case of the disk-infiltrator measurements (Figure 3, Table 2). No distribution of the values between 0 and 3.4 mm h⁻¹ would make the whole distribution log-normal, but it would remain approximately bimodal. A subset of the present DI data has been described as exponential [Nyman *et al.*, 2010]. Most researchers found log-normal distributions of K_s with ponded ring infiltrometers at all scales between 0.05 and 4 m diameter mostly in agricultural soils [Baker and Bouma, 1976; Haws *et al.*, 2004; Lauren *et al.*, 1988; Mallants *et al.*, 1997; Shouse *et al.*, 1994; Sisson and Wierenga, 1981; Wuest, 2005].

The fact that the small scale distributions of K_s are not significantly log-normally distributed contradicts the assumption of the LSC model. There are probably no models where all assumptions are met or tested, and it is a matter of judgment what discrepancies one is willing to accept. In most applications of KINEROS2, for example, the distribution of K_s is neither measured nor calibrated and thus not tested. In this study we chose to use the LSC despite this contradiction, because we would otherwise not be able to achieve our objective of comparing $\overline{K_s}$ across scales with a widely used infiltration model. It would be a limitation of the scope of this study if we restricted ourselves only to test the assumptions of a model, and not the model outcomes.

Coefficients of variation decrease with increased size of the measurement area [Hu *et al.*, 2013; Lai and Ren, 2007; Lauren *et al.*, 1988; Mallants *et al.*, 1997], which we can clearly confirm with our measurements (*CV* ratio in Table 6). This decrease is problematic for the

LSC, because it violates the assumption of a scale-independent CV . When we apply the CV measured at the small scale in the determination of $\overline{K_S}$ at the rainfall simulation or catchment scales, we introduce a dependency between scales, which is difficult to quantify and that was not foreseen in the LSC. Again, as with the log-normality assumption, we chose to proceed with using the LSC to achieve our objective.

● Variations in means across scales have been consistently found for different measurement methods. Pondered infiltration in the field is larger than on smaller cores in the lab or field [Buttle and House, 1997; Davis *et al.*, 1999; Merz *et al.*, 2002]. Most explanations centre on the abundance, functioning, avoidance and cutting off of macropores during measurements [Davis *et al.*, 1999; Mallants *et al.*, 1997; Wuest, 2005], but none of these hypotheses has been tested. Others have found $\overline{K_S}$ values inferred from rainfall simulations were lower by up to an order of magnitude than values from ring infiltrometer measurement [Cerdeira, 1996; Karssenberg, 2006; Merz *et al.*, 2002; Sheridan *et al.*, 2007]. Also, $\overline{K_S}$ values inversely calculated for plots or small watersheds under natural rainfall are consistently lower than values inferred from ring infiltrometers [Chahinian *et al.*, 2006; Williams and Bonell, 1988]. Interestingly, $\overline{K_S}$ values inferred from small watersheds are also smaller than values from rainfall simulations [Burns, 2010; Langhans *et al.*, 2014]. Our results confirm that also for fire-affected soils $\overline{K_S}$ from fully ponded conditions (at the smaller support scale) are higher than $\overline{K_S}$ from partially ponded conditions (at the larger support scale) (Figure 4).

6.2 Accounting for the discrepancy in $\overline{K_S}$ between methods and scales

Authors have offered explanations for this scale-dependency including ‘event-dependency’, rainfall intensity-dependency and spatial heterogeneity (or distribution of infiltration capacities) [e.g. Sheridan *et al.*, 2007; Williams and Bonell, 1988]. Others have modeled or

‘upscaled’ infiltration from the small to the large scale to bridge the gap between measurements, with variable degrees of success [*Chahinian et al.*, 2006; *Karssenberg*, 2006; *Langhans et al.*, 2014; *Merz et al.*, 2002]. Importantly, the latter authors have accounted for rainfall intensity, ‘runon’, and spatial variability in their models. The problem is that all important sub-processes variables that allow up-scaling are: 1) uncertain to be important or dominant, 2) very difficult to measure and therefore rarely comprehensively measured, and 3) likely to be site-specific. The LSC does account for rainfall intensity and spatial heterogeneity, albeit in a spatially abstracted way. This allowed us to avoid the subjective focus on one or more sub-processes that are thought to be important, and rather quantify the overall boundary conditions and distribution effect. Secondly, it allowed us to compare a large data set that was collected with other objectives in mind.

While our study does not provide means to disentangle the components that could explain Figure 4 conclusively, we can offer some cautious interpretations. In fire-affected forest soils the effect of full ponding with unlimited water supply (DI, RI) vs. partial ponding with limited water supply is likely to be of overriding importance. It is very likely that 5 mm water depth (as in DI) is achieved during rainfall simulation and natural rainfall-runoff events over large parts of the soil surface, which makes the methods similar in terms of average water depth. However, for the small-scale methods ponding is full, and water supply unlimited, and water depth therefore unaffected by K_s . But for the larger-scale methods with partial ponding, water depth will be biased towards areas with lower K_s , and water depth or ponding in areas with high K_s will be limited by water supply rates either from rainfall or runoff. In principle, the LSC can account for the effect of full vs. partial ponding, but the observed bias is 4.8 times stronger than predicted by the LSC. We offer the hypothesis that during full ponding with unlimited water supply, the full flow capacity of preferential flow paths connected to the surface is accessed, while during partial ponding with water supply limitation, only a fraction

of this flow capacity is accessed. This water supply limitation would be particularly strong at low aridity sites with high infiltration capacities. Indeed, $\overline{K_S}$ ratios of low aridity (EKW1) are up to an order of magnitude higher than high aridity $\overline{K_S}$ ratios (Table 6). At small scales the actual distribution of hydraulic conductivities is therefore likely to have a large variance. This phenomenon can be attributed to strong water repellency that was reported in the studies used for this analysis [Cawson, 2012; Noske *et al.*, accepted; Nyman *et al.*, 2010; Nyman *et al.*, 2014a; Oono, 2010; Sheridan *et al.*, 2007], and the importance of preferential flow in the region [Burch *et al.*, 1989; Lane *et al.*, 2006; Nyman *et al.*, 2010; Nyman *et al.*, 2014a]. In this context, the partial vs. full ponding effect would be larger in a burnt soil where water repellency is stronger than in unburnt forest: the difference between lower K_s areas (that are water repellent) and areas with higher K_s (preferential flow areas) results in more bias of ponding towards low infiltration areas. Indeed, the $\overline{K_S}$ ratio is higher for burnt than for unburnt sites (Table 6). In addition, some of the water repellent ponding areas might experience activation of water repellent pores through positive water pressure [Ebel *et al.*, 2012; Letey *et al.*, 2000; Nyman *et al.*, 2010]. An alternative explanation of lower infiltration rates with rainfall simulations is that the kinetic energy of the rainfall simulation can alter the transient properties of a recently burnt soil. We think this is unlikely, as all types of measurements were carried out typically after some natural rainfall had already consolidated the surface somewhat, and secondly a decline in infiltration rates during rainfall simulation was not observed.

Within the LSC assumptions, the maximum possible runon effect was quantified by comparing $\overline{K_S}$ calculated with CV set to 0 (no runon possible) and CV as given by the disk or ring infiltrometer measurements (Table 6). The effect was typically between 10 and 25 %, and at most 43 % at the STCN catchment in March 2010. The *actual* runon effect might have been larger, but not so much as to dominate over the partial vs. full ponding effect, or else the

$\overline{K_S}$ ratio would not be so clearly above 1 when comparing disk infiltrometer values with catchment scale values where the runoff effect can be expected to be at its largest.

6.3 Implications of results

In this study we have proven the importance of surface boundary conditions to infiltration and the distribution of K_s on fire-affected hillslopes. We have provided a hypothesis of the dominant process involved, and shown that the scale-independent log-normal distribution assumption might be inapplicable on fire-affected hillslopes: especially at smaller scales the distribution is not log-normal, probably due to water repellency and preferential flow paths.

The study also used a new way to evaluate how a runoff and infiltration model performs in accounting for boundary conditions: by comparing the central parameter that should be constant (e.g. $\overline{K_S}$) across scales and methods, but for the same time and place.

Of course it is widely appreciated that in most soils, (burnt) forest [Nyman *et al.*, 2010], arable [Lauren *et al.*, 1988; Mallants *et al.*, 1997; Wuest, 2005] or grassland [Merz *et al.*, 2002] infiltration is not homogeneous even at the scale of a soil core, but dominated by preferential flow. More generally, heterogeneity is usually organized in patterns that need to be understood for good infiltration modeling [McDonnell *et al.*, 2007]. But there are very few studies that focus on the *interaction* of this soil surface heterogeneity with boundary conditions to infiltration. The three studies mentioned in the introduction are an exception, as they explicitly model water supply to surface connected macropores [Léonard *et al.*, 2001; Ruan and Illangasekare, 1998; Weiler, 2005]. Given the importance of the boundary condition and distribution effect (factor of 4.8 between methods and scales) there should be much more focus on this issue. The dominant process length of this interaction (for example the distribution of flow lengths from infiltration excess to re-infiltration [Sheridan *et al.*,

2014]) is likely to be highly dynamic within and between events depending on boundary conditions.

Despite the complexity, there is no way around designing new models and experiments that take account more explicitly of those interactions to achieve advancement in the problem of runoff generation and scale. As for measurements, it is important to measure both soil infiltration patterns and distribution of conductivities and surface boundary conditions, and to establish the link between them as closely as possible. This is a big challenge and probably requires new experimental methods, such as dye tracing and measuring the distribution of water depth over time. As for models, one focus could be on improving the distribution representation: it should be measurements-based and scale-dependent. Another important recognition is that infiltration rates at field-saturation are not only limited by water supply rates, but that current concepts of K_{fs} (or $\overline{K_S}$, and therefore K_e) itself are dependent on surface boundary conditions.

What does this mean for our purpose of using small-scale measurements for up-scaling runoff and infiltration on hillslopes? It might not be enough to have one type of measurement. One set of measurements should be aimed at characterizing the scale-dependent distribution of hydraulic conductivities at a relatively small scale (e.g. disk infiltrometers). Another one should aim at quantifying the relationship between water input (rainfall and runoff) and infiltration at a slightly larger scale, with additional measurement of surface boundary conditions (e.g. rainfall simulations or rainfall-runoff plots) and resulting infiltration patterns in the soil. From both types, the relationships between hydraulic conductivity, boundary conditions, and infiltration need to be inferred that are at the core of the up-scaling problem.

7. Conclusions

From a large data set of infiltration measurements in burned and unburnt SE Australian forest hillslopes we calculated mean K_s ($\overline{K_s}$) values that were measured at the same time and place, but with different methods and scales. We used a conceptual model of effective hydraulic conductivity (K_e) that accounts for rainfall intensity and assumes a scale-independent log-normal distribution [Smith and Goodrich, 2000]. $\overline{K_s}$ values from ponded disk and ring infiltrometer measurements were on average 4.8 times larger than rainfall simulation and catchment scale values. Coefficients of variation were similarly larger for the small scale compared to the larger scale methods. Furthermore, we rejected log-normality for the disk infiltrometer and catchment scale distributions at a significance level of 0.05 based on the Lilliefors test whereas log-normality for the ring infiltrometer and rainfall simulation values was uncertain, with relatively low p-values of 0.27 and 0.06, respectively. The disk infiltrometer values showed separation into very low values (61 %) and approximately normally distributed larger values. At the disk infiltrometer scale the distribution was very wide (possibly bimodal), probably through the effect of water repellency and preferential flow paths, typical for burnt forest soils. It was hypothesized that during rainfall simulation or natural rainfall-runoff ponding and runoff was strongly biased towards the low conductivity areas whereas the preferential infiltration areas are more water-limited than the K_e concept with its log-normal distribution assumption predicts. On the other hand, the small-scale methods had full ponding with unlimited water supply. This strong discrepancy has been reported before for other environments, and our study underlines the need to improve the measurement and explicit modeling of the interaction of surface boundary conditions and actual distributions of hydraulic conductivities for advancement in scale problems of runoff generation.

Acknowledgements

For accessing the data, please contact the corresponding author at:

christoph.langhans@unimelb.edu.au. The authors would like to thank Melbourne Water for providing research funding and Shea Burns from the USDA-ARS for advice with running KINEROS2 in batch mode.

Accepted Article

References

- Al-Qurashi, A., N. McIntyre, H. Wheeler, and C. Unkrich (2008), Application of the KINEROS2 rainfall-runoff model to an arid catchment in Oman, *Journal of Hydrology*, 355(1-4), 91-105.
- Baker, F. G., and J. Bouma (1976), Variability of hydraulic conductivity in 2 subsurface horizons of 2 silt loam soils, *Soil Science Society of America Journal*, 40(2), 219-222.
- Beven, K. (1989), Changing Ideas in Hydrology - the Case of Physically-Based Models, *Journal of Hydrology*, 105(1-2), 157-172.
- Beven, K., and P. Germann (1982), Macropores and water-flow in soils, *Water Resources Research*, 18(5), 1311-1325.
- Beven, K., and P. Germann (2013), Macropores and water flow in soils revisited, *Water Resources Research*, 49(6), 3071-3092.
- Binley, A., K. Beven, and J. Elgy (1989), A physically based model of heterogeneous hillslopes: 2. Effective hydraulic conductivities, *Water Resour. Res.*, 25(6), 1227-1233.
- Blöschl, G., and M. Sivapalan (1995), Scale issues in hydrological modelling: A review, *Hydrological Processes*, 9, 251-290.
- Budyko, M. I. (1974), *Climate and life*, Academic Press, New York.
- Burch, G. J., I. D. Moore, and J. Burns (1989), Soil hydrophobic effects on infiltration and catchment runoff, *Hydrological Processes*, 3(3), 211-222.
- Burns, I. S. C. (2010), Reducing uncertainty in hydrological modeling: A step-wise multi-scale calibration approach, Ms. thesis, University of Arizona, Tucson, AZ.
- Buttle, J. M., and D. A. House (1997), Spatial variability of saturated hydraulic conductivity in shallow macroporous soils in a forested basin, *Journal of Hydrology*, 203(1-4), 127-142.
- Canfield, H. E., D. C. Goodrich, and I. S. Burns (2005), Selection of parameter values to model post-fire runoff and sediment transport at the watershed scale in southwestern forests, paper

presented at In: Proceedings, ASCE Watershed Management Conference, Williamsburg, VA, July 19-22, 2005.

Cawson, J. (2012), Connectivity of runoff and erosion following planned burns in dry eucalypt forests, PhD thesis, PhD Thesis, University of Melbourne.

Cerda, A. (1996), Seasonal variability of infiltration rates under contrasting slope conditions in southeast Spain, *Geoderma*, 69(3-4), 217-232.

Chahinian, N., M. Voltz, R. Moussa, and G. Trotoux (2006), Assessing the impact of the hydraulic properties of a crusted soil on overland flow modelling at the field scale, *Hydrological Processes*, 20(8), 1701-1722.

Chen, L., M. Berli, and K. Chief (2013), Examining modeling approaches for the rainfall-runoff process in wildfire-affected watersheds: using San Dimas Experimental Forest, *Journal of the American Water Resources Association*, 49(4), 851-866.

Corradini, C., R. Morbidelli, and F. Melone (1998), On the interaction between infiltration and Hortonian runoff, *Journal of Hydrology*, 204(1-4), 52-67.

Davis, S. H., R. A. Vertessy, and R. P. Silberstein (1999), The sensitivity of a catchment model to soil hydraulic properties obtained by using different measurement techniques, *Hydrological Processes*, 13(5), 677-688.

DSE (2012), Simplified Native Vegetation Groups, edited.

Dunne, T., and W. E. Dietrich (1980), Experimental investigation of Horton overland flow on tropical hillslopes; 2) hydraulic characteristics and hillslope hydrographs (Kenya), *Zeitschrift fur Geomorphologie, Supplementband*, 35, 60-80.

Dunne, T., W. H. Zhang, and B. F. Aubry (1991), Effects of Rainfall, Vegetation, and Microtopography on Infiltration and Runoff, *Water Resources Research*, 27(9), 2271-2285.

Ebel, B. A., and J. A. Moody (2013), Rethinking infiltration in wildfire-affected soils, *Hydrological Processes*, 27(10), 1510-1514.

Ebel, B. A., J. A. Moody, and D. A. Martin (2012), Hydrologic conditions controlling runoff generation immediately after wildfire, *Water Resour. Res.*, 48(3), W03529.

Emmett, W. W. (1978), Overland flow, in *Hillslope Hydrology*, edited by M. J. Kirkby, pp. 145-176, Wiley, New York.

Goodrich, D. C., H. E. Canfield, I. S. Burns, D. J. Semmens, S. N. Miller, M. Hernandez, L. R. Levick, D. P. Guertin, and W. G. Kepner (2005), Rapid post-fire hydrologic watershed assessment using the AGWA GIS-based hydrologic modeling tool, paper presented at In: Proceedings of the American Society of Civil Engineers, Watershed Management Conference, Williamsburg, Virginia, July 22-25, 2005.

Goodrich, D. R., C. L. Unkrich, J. R. Smith, and D. A. Woolhiser (2006), KINEROS2 - New features and capabilities, paper presented at Proc. 3rd Fed. Interagency Hydrologic Modeling Conf., April 2-6, 2006, Reno, Nevada.

Grayson, R. B., I. D. Moore, and T. A. McMahon (1992), Physically based hydrologic modeling. 2. Is the concept realistic, *Water Resources Research*, 28(10), 2659-2666.

Green, W. H., and G. A. Ampt (1911), Studies on soil physics, 1. The flow of air and water through soils, *Journal of Agricultural Sciences*, 4(1), 1-22.

Hawkins, R. H. (1982), Interpretation of source area variability in rainfall-runoff relations, paper presented at Proc., Int. Symp. on Rainfall-Runoff Modeling, Mississippi State Univ., Mississippi State, May 18-21, 1981.

Hawkins, R. H., and T. W. Cundy (1987), Steady-state analysis of infiltration and overland-flow for spatially-varied hillslopes, *Water Resources Bulletin*, 23(2), 251-256.

Haws, N. W., B. W. Liu, C. W. Boast, P. S. C. Rao, E. J. Klavivko, and D. P. Franzmeier (2004), Spatial variability and measurement scale of infiltration rate on an agricultural landscape, *Soil Science Society of America Journal*, 68(6), 1818-1826.

- Hu, W., M. G. Shao, Q. J. Wang, and D. L. She (2013), Effects of Measurement Method, Scale, and Landscape Features on Variability of Saturated Hydraulic Conductivity, *Journal of Hydrologic Engineering*, 18(4), 378-386.
- Jones, O. D., G. J. Sheridan, and P. N. Lane (2013), Using queuing theory to describe steady-state runoff-runon phenomena and connectivity under spatially variable conditions, *Water Resources Research*, 49(11), 7487-7497.
- Kalin, L., R. S. Govindaraju, and M. M. Hantush (2003), Effect of geomorphologic resolution on modeling of runoff hydrograph and sedimentograph over small watersheds, *Journal of Hydrology*, 276(1&4), 89-111.
- Karssenbergh, D. (2006), Upscaling of saturated conductivity for Hortonian runoff modelling, *Advances in Water Resources*, 29(5), 735-759.
- Kidwell, M. R., M. A. Weltz, and D. P. Guertin (1997), Estimation of Green-Ampt effective hydraulic conductivity for rangelands, *Journal of Range Management*, 50(3), 290-299.
- Kirchner, J. W. (2006), Getting the right answers for the right reasons: Linking measurements, analyses, and models to advance the science of hydrology, *Water Resources Research*, 42(3), 1-5.
- Lai, J., and L. Ren (2007), Assessing the size dependency of measured hydraulic conductivity using double-ring infiltrometers and numerical simulation, *Soil Science Society of America Journal*, 71(6), 1667-1675.
- Lane, P. N. J., P. B. Hairsine, J. C. Croke, and I. Takken (2006), Quantifying diffuse pathways for overland flow between the roads and streams of the Mountain Ash forests of central Victoria Australia, *Hydrological Processes*, 20(9), 1875-1884.
- Langhans, C., G. Govers, and J. Diels (2013), Development and parameterization of an infiltration model accounting for water depth and rainfall intensity, *Hydrological Processes*, 27(25), 3777-3790.

- Langhans, C., G. Govers, J. Diels, W. Clymans, and A. Van den Putte (2010), Dependence of effective hydraulic conductivity on rainfall intensity: loamy agricultural soils, *Hydrological Processes*, 24(16), 2257–2268.
- Langhans, C., G. Govers, J. Diels, J. J. Stone, and M. A. Nearing (2014), Modeling scale-dependent runoff generation in a small semi-arid watershed accounting for rainfall intensity and water depth, *Advances in Water Resources*, 69(0), 65-78.
- Langhans, C., H. G. Smith, D. M. O. Chong, P. Nyman, P. N. J. Lane, and G. J. Sheridan (2016), A model for assessing water quality risk in catchments prone to wildfire, *Journal of Hydrology*, 534, 407-426.
- Lauren, J. G., R. J. Wagenet, J. Bouma, and J. H. M. Wosten (1988), Variability of Saturated Hydraulic Conductivity in a Glossaquic Hapludalf with Macropores, *Soil Science*, 145(1), 20-28.
- Léonard, J., E. Perrier, and G. de Marsily (2001), A model for simulating the influence of a spatial distribution of large circular macropores on surface runoff, *Water Resources Research*, 37(12), 3217-3225.
- Letey, J., M. L. K. Carrillo, and X. P. Pang (2000), Approaches to characterize the degree of water repellency, *Journal of Hydrology*, 231-232, 61-65.
- Lilliefors, H. W. (1967), On the Kolmogorov-Smirnov Test for Normality with Mean and Variance Unknown, *Journal of the American Statistical Association*, 62(318), 399-402.
- Mallants, D., B. P. Mohanty, A. Vervoort, and J. Feyen (1997), Spatial analysis of saturated hydraulic conductivity in a soil with macropores, *Soil Technology*, 10(2), 115-131.
- McDonnell, J. J., et al. (2007), Moving beyond heterogeneity and process complexity: A new vision for watershed hydrology, *Water Resources Research*, 43(7).
- Mertens, J., D. Jacques, J. Vanderborght, and J. Feyen (2002), Characterisation of the field-saturated hydraulic conductivity on a hillslope: in situ single ring pressure infiltrometer measurements, *Journal of Hydrology*, 263(1–4), 217-229.

- Merz, B., A. Bardossy, and G. R. Schiffler (2002), Different methods for modelling the areal infiltration of a grass field under heavy precipitation, *Hydrological Processes*, 16(7), 1383-1402.
- Morbidelli, R., C. Corradini, and R. S. Govindaraju (2006), A field-scale infiltration model accounting for spatial heterogeneity of rainfall and soil saturated hydraulic conductivity, *Hydrological Processes*, 20(7), 1465-1481.
- Nash, J. E., and J. V. Sutcliffe (1970), River flow forecasting through conceptual models part I -- A discussion of principles, *Journal of Hydrology*, 10(3), 282-290.
- Nearing, M. A., G. R. Foster, L. J. Lane, and S. C. Finkner (1989), A process-based soil-erosion model for USDA-Water Erosion Prediction Project technology, *Transactions of the Asae*, 32(5), 1587-1593.
- Nearing, M. A., B. Y. Liu, L. M. Risse, and X. Zhang (1996), Curve numbers and Green-Ampt effective hydraulic conductivities, *Journal of the American Water Resources Association*, 32, 125-136.
- Noske, P. J., G. Sheridan, P. Nyman, and P. N. J. Lane (accepted), Effects of aridity in controlling the magnitude of runoff and erosion after wildfire, *Water Resources Research*.
- Nyman, P., G. Sheridan, and P. N. J. Lane (2010), Synergistic effects of water repellency and macropore flow on the hydraulic conductivity of a burned forest soil, south-east Australia, *Hydrological Processes*, 24(20), 2871-2887.
- Nyman, P., G. J. Sheridan, H. G. Smith, and P. N. J. Lane (2011), Evidence of debris flow occurrence after wildfire in upland catchments of south-east Australia, *Geomorphology*, 125(3), 383-401.
- Nyman, P., G. J. Sheridan, H. G. Smith, and P. N. J. Lane (2014a), Modeling the effects of surface storage, macropore flow and water repellency on infiltration after wildfire, *Journal of Hydrology*, 513(0), 301-313.

- Nyman, P., C. B. Sherwin, C. Langhans, P. N. J. Lane, and G. Sheridan (2014b), Downscaling regional climate data to calculate the radiative index of dryness in complex terrain, *Australian Metrological and Oceanographic Journal*, 64(2), 109-122.
- Oono, A. (2010), Water repellency and infiltration characteristics of unburnt and burnt Eucalyptus species forests in south-eastern Australia, Msc Thesis, The University of Melbourne.
- Paige, G. B., J. J. Stone, D. P. Guertin, and L. J. Lane (2002), A strip model approach to parameterize a coupled Green-Ampt kinematic wave model, *Journal of the American Water Resources Association*, 38(5), 1363-1377.
- Reynolds, W. D. (1993), Saturated Hydraulic Conductivity: Field Measurements, in *Soil Sampling and Methods of Analysis*, edited by M. R. Carter, pp. 599-613, Lewis Publisher, Florida.
- Reynolds, W. D., and D. E. Elrick (1990), Ponded infiltration from a single ring. 1. Analysis of steady flow, *Soil Science Society of America Journal*, 54(5), 1233-1241.
- Risse, L. M., M. A. Nearing, and X. C. Zhang (1995), Variability in Green-Ampt effective hydraulic conductivity under fallow conditions, *Journal of Hydrology*, 169(1-4), 1-24.
- Ruan, H., and T. H. Illangasekare (1998), A model to couple overland flow and infiltration into macroporous vadose zone, *Journal of Hydrology*, 210(1-4), 116-127.
- Ruan, H. X., L. R. Ahuja, T. R. Green, and J. G. Benjamin (2001), Residue cover and surface-sealing effects on infiltration: Numerical simulations for field applications, *Soil Science Society of America Journal*, 65(3), 853-861.
- Sharma, M. L., G. A. Gander, and C. G. Hunt (1980), Spatial variability of infiltration in a watershed, *Journal of Hydrology*, 45(1-2), 101-122.
- Sheridan, G., P. Nyman, C. Langhans, J. Cawson, P. J. Noske, A. Oono, R. Van Der Sant, and P. N. J. Lane (accepted), Is aridity a high-order control on the hydro-geomorphic response of burned landscapes?, *International Journal of Wildland Fire*.

Sheridan, G. J., P. N. J. Lane, and P. J. Noske (2007), Quantification of hillslope runoff and erosion processes before and after wildfire in a wet Eucalyptus forest, *Journal of Hydrology*, 343(1-2), 12-28.

Sheridan, G. J., P. J. Noske, P. N. J. Lane, O. D. Jones, and C. B. Sherwin (2014), A simple two-parameter model for scaling hillslope surface runoff, *Earth Surface Processes and Landforms*, 39(8), 1049-1061.

Shouse, P. J., T. R. Ellsworth, and J. A. Jobes (1994), Steady-state infiltration as a function of measurement scale, *Soil Science*, 157(3), 129-136.

Sidman, G., D. P. Guertin, D. C. Goodrich, D. Thoma, D. Falk, and I. S. Burns (2015), A coupled modelling approach to assess the effect of fuel treatments on post-wildfire runoff and erosion, *International Journal of Wildland Fire*, -.

Sisson, J. B., and P. J. Wierenga (1981), Spatial variability of steady-state infiltration rates as a stochastic-process, *Soil Science Society of America Journal*, 45(4), 699-704.

Smith, M. W., N. J. Cox, and L. J. Bracken (2010), Modeling depth distributions of overland flows, *Geomorphology*, 125(3), 402-413.

Smith, R. E., and D. C. Goodrich (2000), Model For Rainfall Excess Patterns on Randomly Heterogeneous Areas, *Journal of Hydrologic Engineering*, 5(4), 355-362.

Smith, R. E., D. C. Goodrich, and D. A. Woolhiser (1990), Areal effective infiltration dynamics for runoff of small catchments, paper presented at In: Transactions 14th Int. Congress of Soil Science.

Smith, R. E., D. C. Goodrich, D. A. Woolhiser, and C. L. Unkrich (1995), Kineros - A Kinematic Runoff and Erosion Model, in *Computer models of watershed hydrology*, edited by N. P. Singh, pp. 697-732, Water Resource Publications, Highlands Ranch, CO.

Van den Putte, A., G. Govers, A. Leys, C. Langhans, W. Clymans, and J. Diels (2013), Estimating the parameters of the Green–Ampt infiltration equation from rainfall simulation data: Why simpler is better, *Journal of Hydrology*, 476(0), 332-344.

Weiler, M. (2005), An infiltration model based on flow variability in macropores: development, sensitivity analysis and applications, *Journal of Hydrology*, 310(1-4), 294-315.

Wilcoxon, F. (1945), Individual comparisons by ranking methods, *Biometrics Bulletin*, 1(6), 80-83.

Williams, J., and M. Bonell (1988), The influence of scale of measurement on the spatial and temporal variability of the Philip infiltration parameters - an experimental study in an Australian savannah woodland, *Journal of Hydrology*, 104(1-4), 33-51.

Wood, E. F., M. Sivapalan, and K. Beven (1986), Scale effects in infiltration and runoff production, paper presented at Conjunctive Water Use: Proceedings of the Budapest Symposium, July 1986., IAHS.

Wuest, S. B. (2005), Bias in ponded infiltration estimates due to sample volume and shape, *Vadose Zone Journal*, 4(4), 1183-1190.

Tables

Table 1: Locations and characteristics of the sites where data of at least two different methods in the same time period was available

Site ID	Name	Geographic coordinates	Burn type, burn date	Forest type	Aridity index ¹	Mean slope gradient ²	Experimental methods ³ (data source)	Dates of experimentation
UY	Upper Yarra	37°43'S 145°51'E	Planned burn, April 2009	Dry forest	1.7	0.43	DI, RFS (Cawson, 2012; Oono, 2010)	Jun 2009, May 2010
SUN	Sunday Creek	37°21'S 145°08'E	Wildfire, Feb 2009	Dry forest	2.1	0.53	DI, RFS (Nyman et al., 2014)	Mar 2010
EKW1	East Kiewa	36°48'S 147°12'E	Wildfire, Jan 2003	Wet forest	1.2 – 1.6	0.3 – 0.47	RI, RFS (Sheridan et al., 2007)	Mar, Sep 2005, Feb 2006
EKW2	East Kiewa	36°48'S 147°12'E	Wildfire, Dec 2006	Wet forest	1.2 – 1.6	0.32	DI, RFS (Nyman et al., 2010)	Mar 2007
STCN	Stony Creek North	36°30'S 146°48'E	Wildfire, Feb 2009	Dry forest,	2.5	0.27	DI, CAT (Noske et al., accepted)	Sep 2009, Mar, Sep 2010,
STCS	Stony Creek South	36°30'S 146°48'E	Wildfire, Feb 2009	Dry forest,	2.3	0.27	DI, CAT (Noske et al., accepted)	Sep 2009, Mar, Sep 2010,

¹Aridity index = Net radiation/(latent heat of vaporization*Precipitation), Nyman et al., 2014a

²Mean slope gradient of rainfall simulation sites; STCN and STCS: catchment average slope gradient

³DI: disk infiltrometer; RI: ring infiltrometer; RFS: rainfall simulation; CAT: catchment calibration

Table 2: Summary of methods, their scales, boundary conditions, actual distribution of K_s and the inference method of $\overline{K_S}$ with the Log-normally distributed Strips Concept

Method	Scale (m ²)	Boundary conditions			Distribution	
		Water depth (mm)	Inundation	Water supply rate	Log- normality ^b	Inference of $\overline{K_S}$
Disk infiltrometer (DI)	0.002	5	Full	Unlimited, steady	Reject (p<0.05)	$f_e (= K_e)$
Ring infiltrometer (RI)	0.07	0-60	Full	Unlimited, steady	Uncertain (p=0.27)	$f_e (= K_e)^c$
Rainfall Simulation (RFS)	3	0-10 ^a	Partial	Limited, steady	Uncertain (p=0.06)	$f_e (= K_e)$ and Equation 1
Catchment (CAT)	3000	unknown	Partial	Limited, unsteady	Reject (p<0.05)	KINEROS2

^a Langhans et al, 2013, Dunne and Dietrich, 1980, Smith, 2010, Emmett 1978

^b Lilliefors test (significance level of 0.05)

^c adjusted for water depth, as water depth was substantial on sloping terrain

Table 3: Values of parameters that were fixed in KINEROS2

Parameter	Values
Interception depth	0.5 mm
Canopy fraction	1
G (net capillary drive)	0 mm
$CV(K_s)$	Values given in Table 6
Initial saturation	0.41
Microtopography	Spacing: 10 cm Relief: 2 cm
Water temperature	20° C

Table 4: Nash-Sutcliffe model efficiency (ME) of the modeled hydrographs in the STCN and STCS catchments, and mean and standard deviation of the optimized parameters

Period	Aspect	n events	ME			Optimized values			
			Fraction > 0.5	Fraction < 0.5	Mean > 0.5	Mean $\overline{K_S}$ (mm h ⁻¹)	Std $\overline{K_S}$ (mm h ⁻¹)	Mean Manning' s_n	Std Manning' s_n
Sep 2009	N	36	0.47	0.53	0.85	1.4	0.8	0.14	0.07
	S	30	0.50	0.50	0.85	3.2	1.4	0.18	0.07
Mar 2010	N	18	0.67	0.33	0.86	4.9	6.0	0.18	0.06
	S	14	0.57	0.43	0.87	5.3	0.9	0.23	0.06
Sep 2010	N	21	0.67	0.33	0.84	2.3	1.4	0.18	0.07
	S	18	0.61	0.39	0.84	5.0	3.4	0.21	0.06

Table 5: Rainfall and runoff statistics of the events with ME > 0.5

Period	Aspect	n events	Rainfall event statistics (means)			Runoff statistics (means)		
			Duration (min)	Total (mm)	Intensity I_{1-min} (I_{15-min}) (mm h ⁻¹)	Total (mm)	Peak: max. 1-min (mm h ⁻¹)	Runoff coefficient (mm mm ⁻¹)
Sep 2009	N	17	167	6.8	18.4 (6.9)	3.2	4.4	0.39
	S	15	162	7.2	19.2 (7.4)	2.8	4.2	0.29
Mar 2010	N	12	152	10.5	37.0 (13.1)	4.0	6.8	0.34
	S	8	125	11.0	33.0 (11.7)	3.9	6.5	0.26
Sep 2010	N	14	358	14.4	26.6 (9.6)	6.8	5.8	0.38
	S	11	471	20.7	33.8 (13.5)	8.5	8.2	0.36

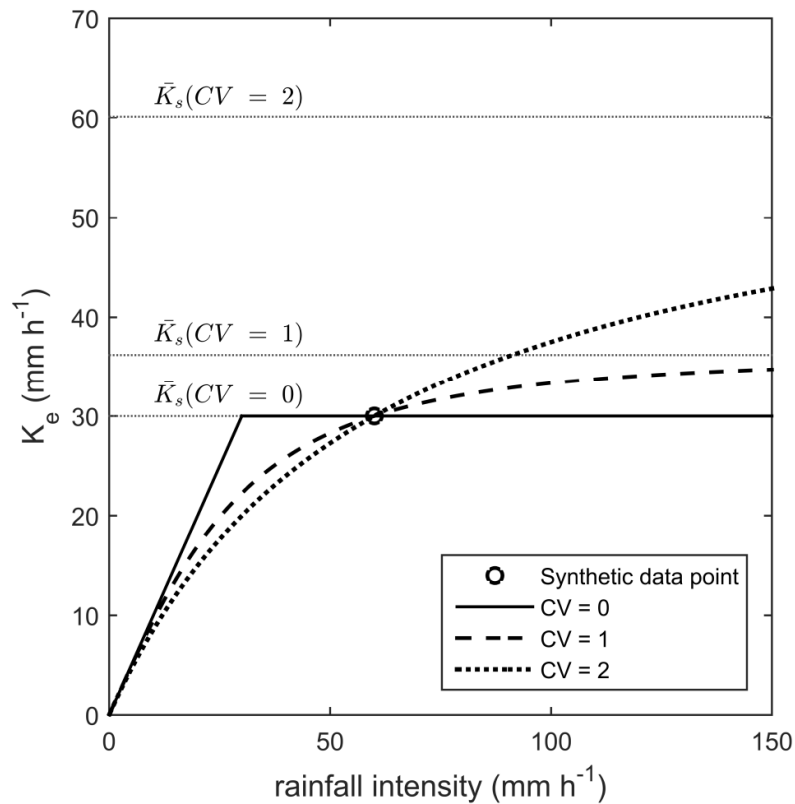
Table 6: Descriptive statistics for all data, separated by site, period, burn condition and method. The $\overline{K_s}$ and CV ratios compare the smaller-scale over the larger-scale value

Site ID (Aridity Index)	Date	Burn condition	Method ^a	<i>n</i>	Mean $\overline{K_s}$ (CV = 0)	Mean $\overline{K_s}$ (CV ^b)	CV	$\overline{K_s}$ ratio ^c	CV ratio ^d		
					mm h ⁻¹						
UY (1.7)	Jun 2009	Burnt	DI	147	30.6	30.6	2.66	*5.3	4.3		
		Burnt	RFS	8	4.4	6.0	0.62				
		Unburnt	DI	45	78.5	78.5	1.91				
		Unburnt	RFS	4	19.2	28.9	0.49				
	Mar 2010	Burnt	DI	48	81.4	81.4	1.30	5.1	4.0		
		Burnt	RFS	3	14.5	16.0	0.32				
		Unburnt	DI	24	52.7	52.7	1.65			1.6	4.5
		Unburnt	RFS	4	26.2	36.2	0.36				
SUN (2.1)	Mar 2010	Burnt	DI	36	46.8	46.8	2.27	4.9	3.6		
		Burnt	RFS	3	8.0	9.9	0.62				
EKW1 (1.4)	Mar 2005	Burnt	RI	6	1350.5	1350.5	0.66	*17.4	1.5		
		Burnt	RFS	3	64.2	98.7	0.45				
		Unburnt	RI	6	732.2	732.2	0.16			*8.6	14.9
		Unburnt	RFS	3	82.4	86.3	0.01				
	Sep 2005	Burnt	RI	6	439.1	439.1	0.45	*4.2	12.4		
		Burnt	RFS	3	88.9	144.9	0.04				
		Unburnt	RI	2	476.0	476.0	0.62			4.2	5.4
		Unburnt	RFS	3	89.9	228.5	0.11				
	Feb 2006	Burnt	RI	6	354.6	354.6	0.80	*9.9	2.1		
		Burnt	RFS	3	32.8	36.0	0.38				
		Unburnt	RI	6	248.7	248.7	0.91			*3.7	7.5
		Unburnt	RFS	3	56.8	74.5	0.12				
EKW2 (1.4)	Mar 2007	Burnt	DI	29	118.1	118.1	1.75	2.7	6.0		
		Burnt	RFS	3	33.9	49.0	0.29				
STCN (2.5)	Sep 2009	Burnt	DI	24	1.3	1.3	0.00	*0.9	0.0		
		Burnt	CAT	17	1.4	1.4	0.55				
	Mar 2010	Burnt	DI	28	20.5	20.5	1.93	4.2	1.6		
		Burnt	CAT	12	2.8	4.9	1.23				
	Sep 2010	Burnt	DI	24	6.1	6.1	1.49	2.7	2.5		
		Burnt	CAT	14	2.1	2.3	0.60				
STCS (2.3)	Sep 2009	Burnt	DI	22	0.0	0.0	0.00	*0.0	0.0		
		Burnt	CAT	16	3.0	3.0	0.49				
	Mar 2010	Burnt	DI	26	32.0	32.0	1.21	6.1	7.4		
		Burnt	CAT	8	4.7	5.3	0.16				
	Sep 2010	Burnt	DI	22	7.6	7.6	1.47	1.5	2.2		
		Burnt	CAT	11	4.2	5.0	0.68				

^aDI: disk infiltrometer; RI: ring infiltrometer; RFS: rainfall simulation; CAT: catchment calibration

^bCoefficient of variation of DI or RI used to calculate $\overline{K_s}$ of RFS and CAT; ^cEquation 4a; ^dEquation 4b

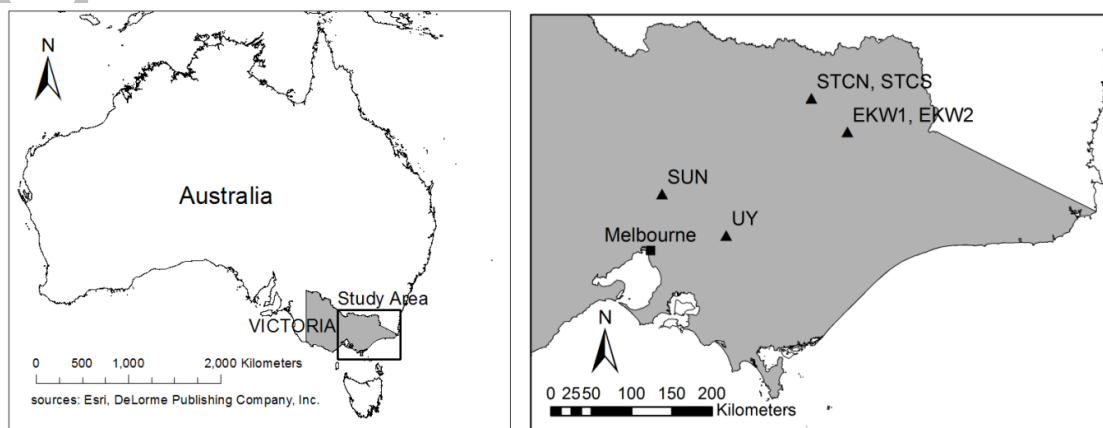
*Median values of small and large scale methods are significantly different, based on the Wilcoxon ranking test (alpha = 0.05)



855

856 Figure 1: Demonstration how K_e varies as a function of rainfall intensity at different CV . Given a data
 857 point, for example derived from steady state infiltration rate of a rainfall simulation, the assumption of
 858 a certain value of CV strongly influences the inferred $\bar{K}_s$ value (dotted lines).

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861 Figure 2: Study site locations in Victoria, SE Australia

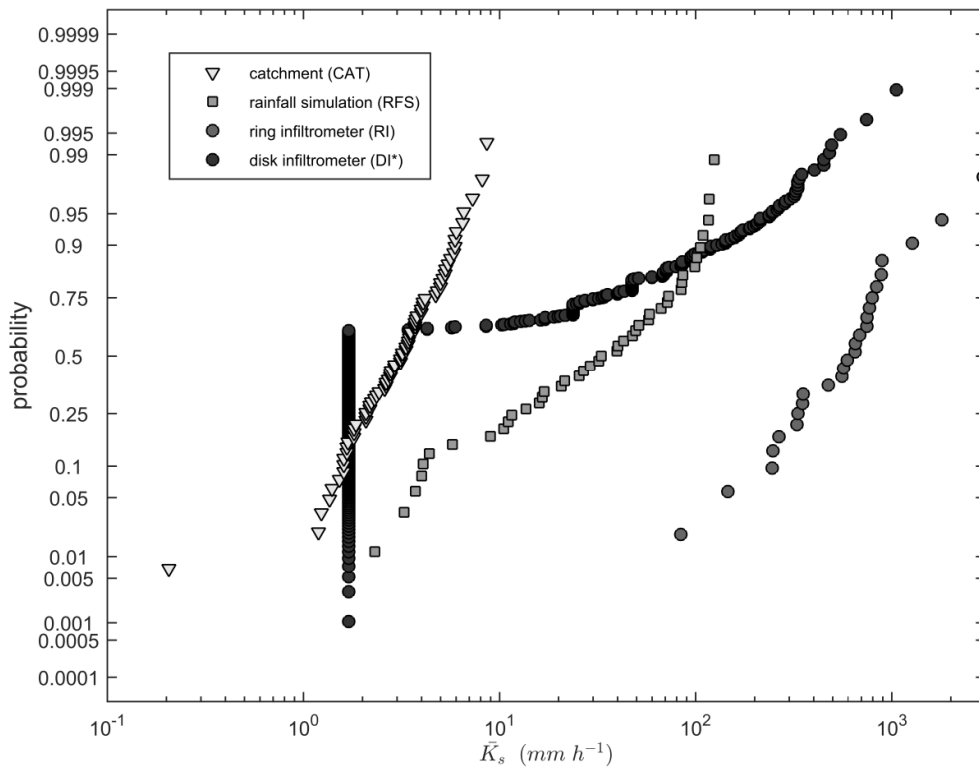


Figure 3: Cumulative probability for $\overline{K}_S$ values of a method. * 61% of disk infiltrometer measurements had values that were below the reading threshold for the method of 3.4 mm h^{-1} . These values are displayed as 1.7 mm h^{-1} (forming a vertical line).

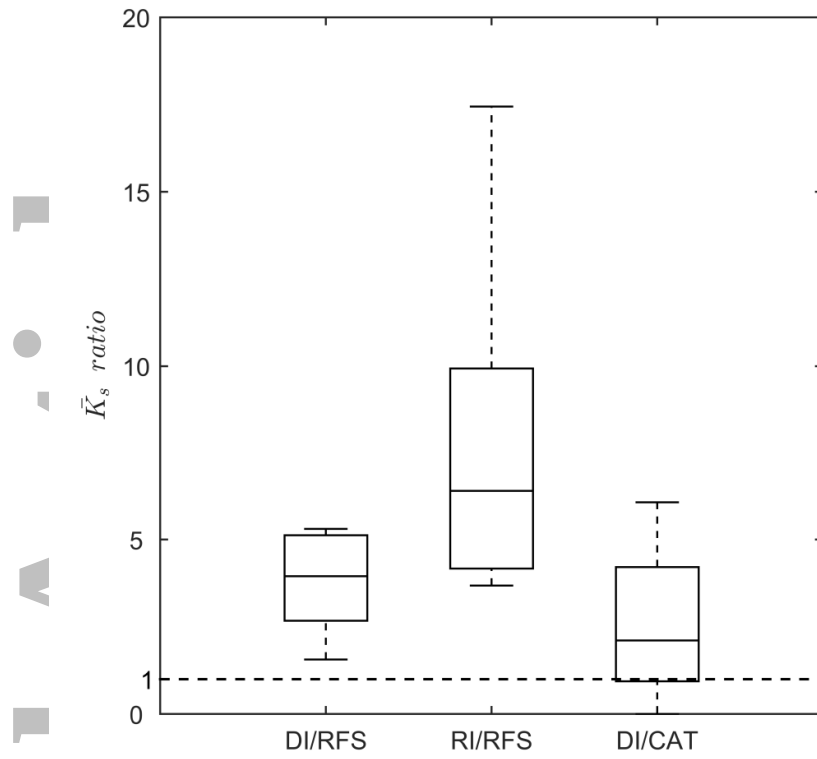


Figure 4: Ratio of smaller-scale (fully ponded) $\bar{K}_S$ values over comparable larger-scale (partially ponded) $\bar{K}_S$ values, for each measured combination of small and large scale measurements at one time and place (Equation 4a)

